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Final Report

Quantification of the Spatial Variability of the Ocean Surface Roughness and Momentum Flux

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ABSTRACT

It is well known that physical processes occurring at the air-sea interface, such as air-sea mass, momentum and energy exchanges, are controlled by the ocean surface roughness that has often been characterized in the literature as being homogeneous in space and time from a large-scale point of view. However, the ocean surface roughness is heterogeneous within a wavelength. The surface areas on wave crests are usually rougher than those on wave troughs. The mechanisms range from capillary ripples on wave crests to larger scale wave breaking. Ambient swell and non-uniform currents can also cause significant spatial variations of the surface roughness as short waves become steeper on the swell crest than those on the swell trough. The study parameterizes the spatial variability of the sea surface roughness and quantifies the resultant variations of momentum and energy exchanges between the atmosphere and the coastal ocean using a phase-resolving numerical wave model. Results have been presented at conferences and submitted to a journal for possible publication.

TABLE OF CONTENTS

1.0 INTRODUCTION

2.0 HETEROGENEOUS AIR-SEA MOMENTUM FLUX

2.1 Parameterization of Momentum Flux

2.2 Effects of Wave Nonlinearity on the Drag Coefficient

3.0 AN ADVANCED BOUSSINESQ WAVE MODEL WITH WIND FORCING

4.0 MODELING WAVE GROWTH ON A SHALLOW LAKE

5.0 APPLICATIONS

6.0 SUMMARY AND CONCLUSIONS

7.0 ACKNOWLEDGEMENTS

8.0 REFERENCES

9.0 APPENDIX

Figures

1.0 INTRODUCTION

Many naval operations occur over and in the atmosphere-ocean interface. The success of such operations depends on the understanding of the structure and behavior of the air-sea interface. It is well known that many physical processes occurring at the air-sea interface, such as air-sea mass, momentum and energy exchanges, are controlled by the ocean surface roughness that has often been characterized in the literature as being homogeneous in space and time from a large-scale point of view. Recently, it is realized, however, that the utilization of remote sensing techniques for the study of the air-sea interface requires the in-depth understanding of the spatial and temporal variability of the ocean surface roughness at multiple scales.

The ocean surface roughness is indeed inhomogeneous within a wavelength. The surface areas on wave crests are usually rougher than those on wave troughs. The mechanisms range from capillary ripples on wave crests to larger scale wave breaking (see e.g., Longuet-Higgins, 1963; Jiang et al., 1999; Ducan et al., 1999). An ambient swell or current can also cause significant spatial variations of the surface roughness as short waves become steeper on the swell crest than those on the swell trough (Longuet-Higgins and Stewart, 1960). The purpose of the proposed study is to quantify the spatial variability of the ocean surface roughness and the resultant variations of momentum and energy exchanges between the atmosphere and the ocean using a phase-resolving numerical wave model.

The report is organized as follows. First, we present the new parameterization of the momentum flux transferred from the wind to surface gravity waves in the coastal zone. Next, we analyze the effect of wave nonlinearity on the drag coefficient using Stokes' second-order wave theory. After that, we show how to implement the parameterized wind stress into the phase resolving Boussinesq wave model with enhanced dispersion properties. Then tests of the extended Boussinesq model with the wind forcing against wave measurements in a shallow lake are carried out followed by an application of the model to the study of sea-swell interactions. Finally, we summarize the findings of the project.

2.0 HETEROGENEOUS AIR-SEA MOMENTUM FLUX

2.1 Parameterization of Momentum Flux

The momentum flux transferred from winds to surface waves in the time-domain Boussinesq wave model may be parameterized as the wind stress, $\tau = \rho_a C_d \mathbf{U} |\mathbf{U}|$, where \mathbf{U} is the wind velocity vector at a reference elevation, and C_d is the corresponding drag coefficient. Suggested by the field evidence, the wind drag coefficient in the coastal zone must be a function of not only the wind speed, but also the geometry of the surface gravity waves that are often skewed and asymmetrical owing to the seabed boundary. We propose a new formula of drag coefficient as follows.

$$C_d 10^3 = (a_1 + a_2 |\eta_x|) + b U_{10} \quad (1)$$

where a_1 , a_2 and b are empirical coefficients, and η_x is the instantaneous surface slope computed in the Boussinesq model. Phase average of the drag coefficient allows for a comparison of the new

formula with conventional drag coefficient formulas, such as the equation proposed by Wu (1980).

Figure 1 illustrates the phase-averaged drag coefficient as a function of wind speed and wave steepness, ka (k = wave number and a = wave amplitude). We choose $a_1=0.2$, $a_2=18$, and $b=0.065$. The wave steepness ranges from 0.05 to 0.2 with an increment of 0.05. It is seen that the new drag coefficient increases with both wind speed and wave steepness. Following Wu (1980), the third term in Equation (1) is set to be 0.4875 for wind speed less than 7.5m/s. The dashed line in Figure 1 depicts the drag coefficient given by Wu

$$C_d 10^3 = 0.8 + 0.065U_{10} \quad (2)$$

which is only a function of wind speed. The curves of the new drag coefficient are parallel to Wu's curve and the offset depends on the wave steepness.

In addition to Wu (1980)'s equation, a large number of empirical formulas for the drag coefficient have been proposed in the literature (see e.g. Geernaert, 1990). Each formula represents the best fit to a specific dataset. Figure 2 shows a comparison of the new drag coefficient and the existing formulations. It is seen that the scattering of the drag coefficient as a function of the wind speed may be explained by the effect of the sea state, or the wave steepness. Statistically, Equation (1) agrees fairly well with all the data, which is not achieved by any single existing formula of drag coefficient.

One of the crucial differences in airflow over steep surface waves in comparison to the flow over a flat surface is the presence of a form drag due to the separation of airflow. Laboratory measurements of the tangential and normal stresses on the air-water interface (Banner and Peirson, 1998) suggest that the form drag accounts for the major proportion of the wind stress once the waves have developed beyond their early growth stage. The most interesting scenario of wind effects on coastal waves is the strong wind condition. We hypothesize that the separation of airflow occurs on the lee side of the wave crest in the shallow water and, consequently, the form drag is a dominant contributor to the momentum flux between the atmosphere and the nearshore wave field. The hypothesis is in line with the sheltering theory proposed by Jeffreys (1925) as a mechanism of wave generation.

The time-domain Boussinesq model resolves the individual wave motion and predicts reasonably well the geometry of the surface waves distorted by the seabed, such as the skewness and asymmetry of the waves in the shallow water (e.g. Chen et al., 2000 and Shi et al. 2003). The wind stress implemented into the Boussinesq model should vary over a wavelength, with a larger drag coefficient on the wave crest than that in the trough to take the form drag into account. In fact, only the heterogeneous distribution of the wind stress over a wavelength allows for changes of wave height by a wind in a phase-resolving wave model. Assuming that the form drag is dominant over the skin friction, we apply the wind stress on the wave crest only, neglecting the effect of the shear stress on the wave trough in the Boussinesq model.

The temporal and spatial variation of drag coefficient is one of the important features of the wind stress on the nearshore waves simulated by the phase-resolving Boussinesq model. With the instantaneous wave celerity and wave slope obtained from the computed surface elevation at each grid and every time step, the wind stress is calculated as a function of the wind velocity relative to the

wave celerity as well as the wave slope in the following form

$$\tau = C_d \rho_a |U_{10} - C| (U_{10} - C) \quad (3)$$

where C is the instantaneous wave celerity estimated in the Boussinesq model. Equation (3) is similar to the wind stress formulation followed by Schwab et al. (1984) in a wave prediction model on the basis of a phase-averaged momentum balance equation rather than an energy balance equation.

2.2 Effects of Wave Nonlinearity on the Drag Coefficient

The new formulation of wind drag coefficient introduced in the preceding section depends on the wave steepness and becomes effective only on the wave crest. It is well known that the effects of wave nonlinearity increase the wave steepness and result in peaky wave profiles. We shall analyze the effect of wave nonlinearity on the drag coefficient given by Equation (1) using Stokes' second-order wave theory.

The free surface profile given by Stokes' second-order theory reads

$$\eta = a \cos(kx - \omega t) + \frac{a^2 k (2 + \cosh 2kh) \cosh kh}{4 \sinh^3 kh} \cos 2(kx - \omega t) \quad (4)$$

where a is the first-order (linear) amplitude of the free-surface disturbance, $k = \frac{2\pi}{L}$ is the wave number, h is the still water depth, L is the wave length, x is the horizontal distance, t is time, and ω is the angular frequency. At $t = 0$ and in the deep water ($h \gg L/2$), Equation (4) may be simplified as

$$\eta = a \cos kx + \frac{a^2 k}{2} \cos 2kx \quad (5)$$

Upon the use of the trigonometric identity and rearrangement of Equation (5), we obtain

$$\eta k = a^2 k^2 \cos^2 kx + ak \cos kx - \frac{a^2 k^2}{2} \quad (6)$$

Because the wind stress (or drag coefficient) is only applied on the wave crest above the still water level, we need to find out the length of the wave crest. Solving Equation (6) with $\eta k = 0$ yields

$$x_0 k = \arccos[f(ak)] \quad (7)$$

in which

$$f(ak) = -\frac{1}{2ak} + \sqrt{\left(\frac{1}{2ak}\right)^2 + \frac{1}{2}} \quad (8)$$

Integrating the drag coefficient over a wave length leads to the phase-averaged drag coefficient

$$\bar{C}_d = \frac{1}{L} \int_{L-x_0}^{L+x_0} C_d dx = \frac{a_1}{\pi} kx_0 + \frac{a_2}{\pi} k\eta_{\max} + \frac{b}{\pi} kx_0 U_{10} \quad (9)$$

where η_{\max} is the surface elevation at the wave crest. For linear waves, we have $x_0 = \frac{L}{4}$ and $\eta_{\max} = a$.

The ratio of the drag coefficient on a Stokes' second-order wave profile to that on a linear wave profile is given by

$$\frac{\bar{C}_d}{(\bar{C}_d)_{\text{linear}}} = 1 + \frac{(a_1 + bU_{10})(2kx_0 - \pi) + a_2(ka)^2}{(a_1 + bU_{10})\pi + 2a_2ka} \quad (10)$$

Figure 3 illustrates the variation of such a ratio as a function of wave steepness and wind speed. It is seen that the wave nonlinearity can increase the phase-averaged drag coefficient by 10-15 percent.

AN ADVANCED BOUSSINESQ WAVE MODEL WITH WIND FORCING

The formulation of the wind stress is implemented into the Boussinesq wave model developed by Wei et al. (1995), Kennedy et al. (2000), Chen et al (2000), and Chen et al. (2003a). In order to take into account the short waves generated by local winds, we also extend the dispersion accuracy of the one-dimensional version of the model to $kh = 6$ (h = still water depth) by introducing the additional terms from Madsen and Schaffer's (1998) equations into the model. A new set of coefficients for better shoaling properties of the equations (Madsen, 2002, personal communication, and Kennedy et al, 2002) is also implemented into the model. In one horizontal dimension, the extended Boussinesq equations read

$$\eta_t + M_x = 0 \quad (11)$$

where

$$M = (h + \eta) \left[u_\alpha + \left(\frac{z_\alpha^2}{2} - \frac{1}{6} (h^2 - h\eta + \eta^2) \right) u_{\alpha\alpha} + \left(z_\alpha + \frac{1}{2} (h - \eta) \right) (hu_\alpha)_{xx} \right] \\ + (\beta_2 - \beta_1) h^2 ((h + \eta) u_\alpha)_{xx} - \beta_2 (h^2 ((h + \eta) u_\alpha)_x)_x + (\beta_2 - \beta_1) h^2 \eta_{xt} - \beta_2 (h^2 \eta_t)_x \quad (12)$$

and

$$u_{\alpha t} + u_\alpha u_{\alpha\alpha} + g\eta_x + \Lambda_0 + \Lambda_1 + \Lambda_2 + \Lambda_3 - R_b - R_s + R_f - R_w = 0 \quad (13)$$

where

$$\Lambda_0 = \left(\frac{z_\alpha^2}{2} + (\alpha_2 - \alpha_1) h^2 \right) u_{\alpha\alpha t} + (z_\alpha - \alpha_2 h) (hu_\alpha)_{xx} + (\alpha_2 - \alpha_1) h^2 \eta_{xxx} - \alpha_2 h (h\eta_x)_{xx} \quad (14)$$

$$\Lambda_1 = \left[z_\alpha u_\alpha (hu_\alpha)_{xx} + \frac{1}{2} z_\alpha^2 u_\alpha u_{\alpha\alpha x} - \eta (hu_\alpha)_x + \frac{1}{2} ((hu_\alpha)_x)^2 \right]_x + \frac{1}{2} (\alpha_2 - \alpha_1) h^2 (u_\alpha^2)_{xxx} - \frac{1}{2} \alpha_2 h (h(u_\alpha^2)_x)_{xx} \quad (15)$$

$$\Lambda_2 = \left[\eta u_{\alpha x} (hu_\alpha)_x - \eta u_\alpha (hu_\alpha)_{xx} - \frac{1}{2} \eta^2 u_{\alpha x t} \right]_x \quad (16)$$

$$\Lambda_3 = \left[\frac{1}{2} \eta^2 (u_{\alpha x})^2 - \frac{1}{2} \eta^2 u_\alpha u_{\alpha\alpha x} \right]_x \quad (17)$$

In the equations, η is the free surface elevation relative to the still water level; u_α is the velocity at the reference elevation $z_\alpha (= -0.54122h)$ in the water column; and the subscripts t and x denote time and spatial differentiations, respectively. The dispersion enhancement coefficients with improved shoaling properties are $(\beta_1, \beta_2) = (0.03917, 0.315236)$ and $(\alpha_1, \alpha_2) = (0.01052, 0.124537)$. The additional terms, R_b , R_s , R_f , and R_w represent the effects of wave breaking, subgrid lateral turbulent mixing, seabed shear stress, and wind stress, respectively. Detailed descriptions of the first three terms can be found in Chen et al. (1999). Upon the use of Equation (3), we obtain

$$R_w = \frac{\rho_a}{\rho(h+\eta)} C_d |U_{10} - C| (U_{10} - C) \quad (18)$$

where ρ is the water density.

The internal wave-maker for the generation of random waves at the offshore boundary is extended to accommodate the enhancement of dispersion properties in the Boussinesq equations. In comparison to Wei et al.'s (1995) model, the additional terms in Equation (12) requires the solution of a tridiagonal system for the continuity equation.

Other extensions of the model include the modification of the breaking criteria in the model to account for the energy dissipation owing to whitecaps. The original breaking criteria in Kennedy et al. (2000) were designed for depth-limited wave breaking. The basic idea is to replace the still water depth h by C^2/g where g is the gravitational acceleration. This leads to the modified criteria for the onset ($\eta_i^{(I)}$) and cessation ($\eta_i^{(F)}$) of wave breaking as follows

$$\begin{aligned} \eta_i^{(I)} &= \gamma_1 C \\ \eta_i^{(F)} &= \gamma_2 C \end{aligned} \quad (19)$$

in which $\gamma_1 = 0.35 \sim 0.65$ and $\gamma_2 = 0.15$. We choose the lower limit of γ_1 in the presence of a wind that is in the direction of the wave train. Other terms with h in the breaking scheme of Kennedy et al. (2000) are also modified in a similar fashion.

4.0 MODELING WAVE GROWTH ON A SHALLOW LAKE

The test cases in Chen et al. (2002) were focused on the surface roughness length and drag coefficient associated with shoaling waves over a distance of about 500m. Can the wind stress implemented into the Boussinesq model simulate the growth of wind waves over a longer fetch? We shall address such a question using the data set of wave growth in Lake George, Australia collected by Young and Verhagen (1996).

Lake George is a shallow lake with a typical water depth of 2m. It is about 20km long and 10km wide. A series of eight observation stations were deployed along the north-south fetch to measure the wind waves. The wave data collected by Young and Verhagen (1996) under a nearly ideal condition have served as a test bed for a number of phase-averaged wave models, such as SWAN (Simulating Waves Nearshore, Booij et al., 1999). We choose two data sets corresponding to a medium wind speed of $U_{10} = 10.8\text{m/s}$ and a large wind speed of $U_{10} = 15.2\text{m/s}$ to test the wind stress formulation incorporated into the Boussinesq model. Owing to the dispersion limit of the Boussinesq model ($kh < 6$), only the observations at the last four stations (5-8) are utilized. The fetch between Stations 5 and 8 is about 4.25 km. The up-wave boundary condition is taken from the observation at Station 5 where the zero-moment wave height is $H_{m0} = 0.367\text{m}$ and the peak wave period is $T_p = 2.34\text{s}$ in the case of a moderate wind ($U_{10} = 10.8\text{m/s}$), and $H_{m0} = 0.473\text{m}$ and $T_p = 2.4\text{s}$ in the case of a strong wind ($U_{10} = 15.2\text{m/s}$). We use *TMA* shallow water wave spectra (Bouws et al., 1985) with the shape parameter $\gamma = 3.3$ at the up-wave boundary as the input to the Boussinesq model. A bottom friction coefficient of $f = 0.0005$ is used in the quadratic law of the bottom shear stress (Chen et al., 1999).

Figure 4 shows comparisons of the computed and measured significant wave heights as well as the computed and inferred drag coefficients under the moderate wind condition. It is seen that the wave heights predicted by the Boussinesq model are in fairly good agreement with the field measurements. The computed peak wave periods, however, do not agree with the observations. The wind stress incorporated into the Boussinesq model simply does not lead to the increase of wave period along the fetch. The downshift of the peak frequency is absent in the modeled wave energy spectrum. This is attributed to the inability of the one-dimensional Boussinesq model to take quadruplet interactions into account. Interestingly, the phase-averaged drag coefficient calculated from the Boussinesq model is in good agreement with the coefficient estimated using the modeled wave characteristics and Anctil and Donelan's (1996) formula. The drag coefficients given by Taylor and Yelland's (2001) formula (triangles), however, are smaller than either Boussinesq or Anctil and Donelan's result (solid line and squares).

Similar comparisons are made for the case of large wind speed as shown in Figure 5. Again, the Boussinesq model predicts the growth of wave height fairly well in comparison to the field measurements but considerably under-predicts the increase of wave period. The agreement among the drag coefficients is also very similar to the case with a moderate wind. The phase-averaged drag coefficient obtained from the Boussinesq model agrees better with Anctil and Donelan's result than does Taylor and Yelland's formula, as shown in Figure 5b. Notice that all the drag coefficients are computed using the wave field given by the one-dimensional Boussinesq model, which is not able to predict the downshift of the peak frequency. Nevertheless, the prediction of wave height growth under

both moderate and strong wind conditions by the Boussinesq model confirms that the new formulation of wind stress is a good representation of the momentum flux transferred from a wind to surface waves in a phase-resolving model.

5.0 APPLICATIONS

Donelan (1987) among others has found that the addition of mechanically-generated swell to the wind waves in a wave flume results in a pronounced reduction in the energy of the wind sea while the swell gains significant energy from the following wind. Although a number of hypotheses have been proposed in the literature, the attenuation mechanism for the wind sea owing in the presence of swell is still poorly understood.

On the basis of the Boussinesq model incorporating the wind forcing, we seek answers to the following questions: 1) Can the mechanism of sea/swell interactions (Longuet-Higgins and Stewart, 1960) explain the spatial variations of sea surface slope computed by the Boussinesq model? 2) How does swell influence spatial variations of the sea surface slope of a random wind sea? 3) What is the ratio of the mean square slope on the swell crest to that on the swell trough? 4) Is the heterogeneous surface roughness able to explain the reduction of the wind-wave growth rate caused by the swell?

As demonstrated in the previous section, the extended Boussinesq model is able to simulate the growth of wave energy for a given wind speed and a non-zero wave spectrum at the up-wave boundary. We use the model with the wind forcing to simulate two laboratory experiments of sea/swell interactions conducted by Hatori et al. (1981) and Donelan (1987). The modeling effort serves two purposes. First, the parameterization of the wind stress in the phase-resolving Boussinesq model is tested against the laboratory observation with respect to the growth of wind waves in the presence of mechanically-generated swell. Second, we utilize the model to explore the connection between the reduced growth rate of wind waves by the presence of ambient swell and the spatial variations of the free surface roughness owing to sea/swell interactions. For test cases beyond the model limit of dispersion accuracy (the dimensionless wave number $kh > 6$), qualitative, rather than quantitative, comparisons with the laboratory observations are made.

Based on the numerical results given by the extended Boussinesq model with the wind forcing, surface slopes, curvatures, and the wind stress or the momentum transfer from the wind are computed. Then the following hypothesis is tested: The spatial variation of ocean surface roughness owing to the presence of swell results in the spatial variation of momentum transfer from the wind and consequently slows the growth of the wind waves.

We expect that on the swell crests, the larger surface roughness leads to a larger momentum transfer than that on the swell trough, but the momentum/energy flux averaged over the swell wavelength remains relatively unchanged. It is also anticipated that the swell takes advantage of the stronger momentum flux on its crest and harvests most of the energy transferred from the wind to the water. In plain words, the finite amplitude swell serves as a "master" wave and the wind waves riding on the swell act as "slaves" for momentum transfer. Consequently, the swell amplitude increases while the wind sea barely grows.

The study sheds light on the relationship between the slow wind wave growth rate in the presence of swell and the spatial variations of the ocean surface roughness. The result is useful for the improvement of phase-averaged wave prediction models that fail to satisfactorily take sea-swell interactions into account. Results will be presented at the 29th International Conference on Coastal Engineering.

6.0 SUMMARY AND CONCLUSIONS

A new wind stress formula that takes the spatial variation of surface roughness over a wave length into account has been developed and analyzed. The report documents (1) the new parameterization of the momentum flux transferred from the wind to surface gravity waves in the coastal zone, (2) the analysis of the effect of wave nonlinearity on the drag coefficient based on Stokes' second-order wave theory, (3) the implementation of the parameterized wind stress into a phase resolving Boussinesq wave model with enhanced dispersion properties, (4) tests of the extended Boussinesq model with the wind forcing against wave measurements in a shallow lake. Fairly agreement between the model results and field data has been found. Results of the study have been submitted to the ASCE Journal of Waterway, Port, coastal and Ocean Engineering for possible publication (Chen et al. 2003b).

The methodology for the parameterization of the air-sea momentum flux as well as the extended Boussinesq model incorporating the wind effects appear to be a promising tool for the study of sea-swell-wind interactions, wind effects on nearshore wave propagation and horizontal circulation, and the spatial variability of the sea surface roughness length. Preliminary results were presented at the 28th International Conference on Coastal Engineering (Chen et al. 2002). An application of the model to the study of swell-sea interactions will be presented at the 29th International Conference on Coastal Engineering (Chen et al. 2003c).

7.0 ACKNOWLEDGEMENTS

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9.0 APPENDIX

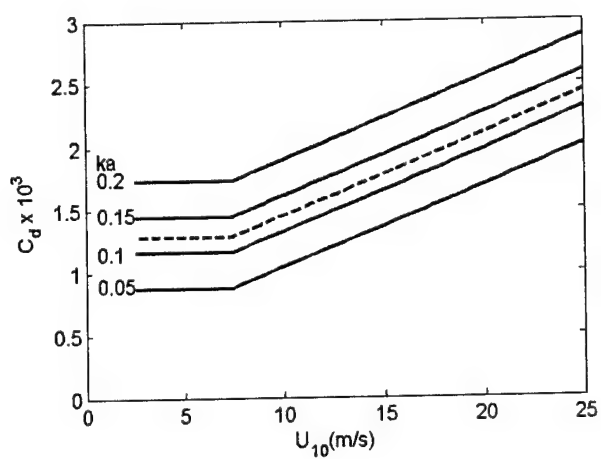


Fig. 1 Drag coefficient as a function of wind speed and wave steepness (solid lines). The dashed line is Wu's (1980) drag coefficient.

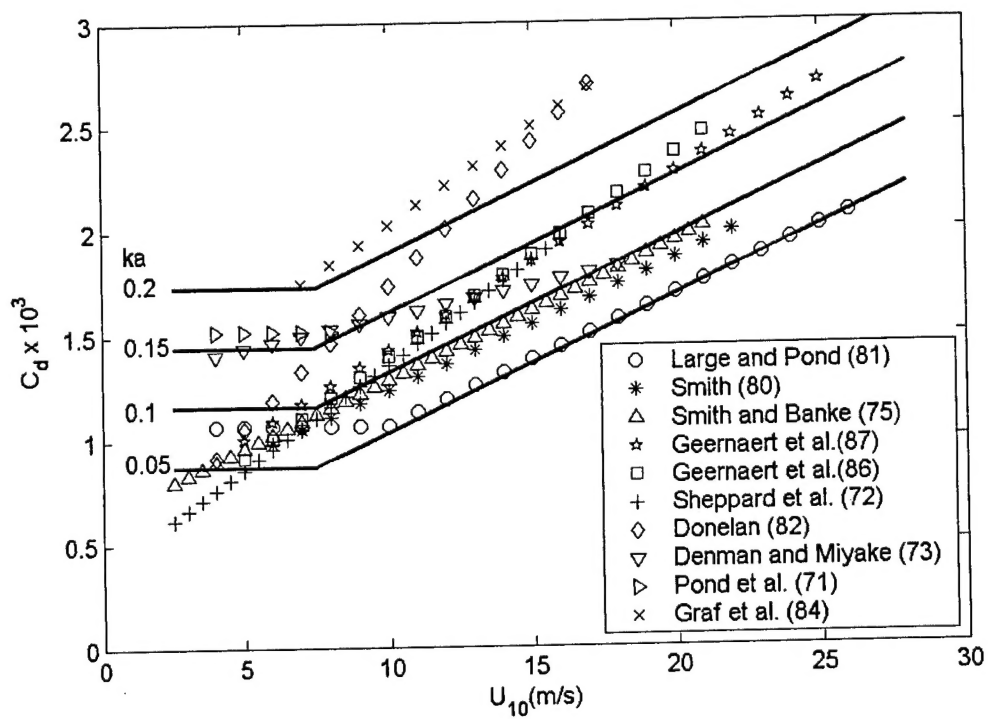


Fig. 2 Comparison of Equation (1) with the drag coefficient formulas compiled by Geernaert (1990).

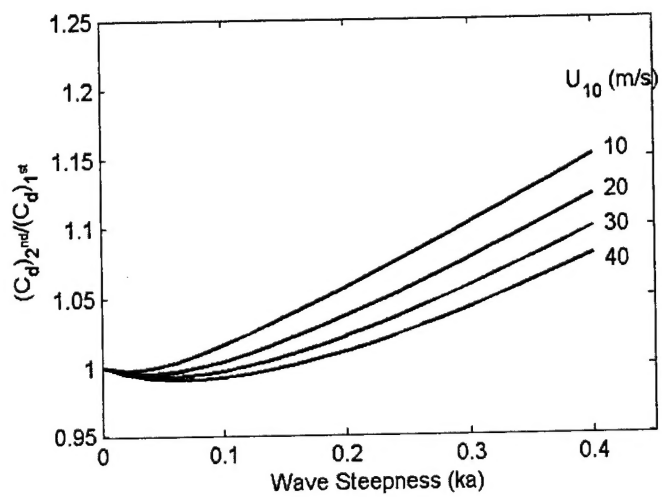


Fig. 3 Effects of wave steepness and wind speed on the drag coefficient

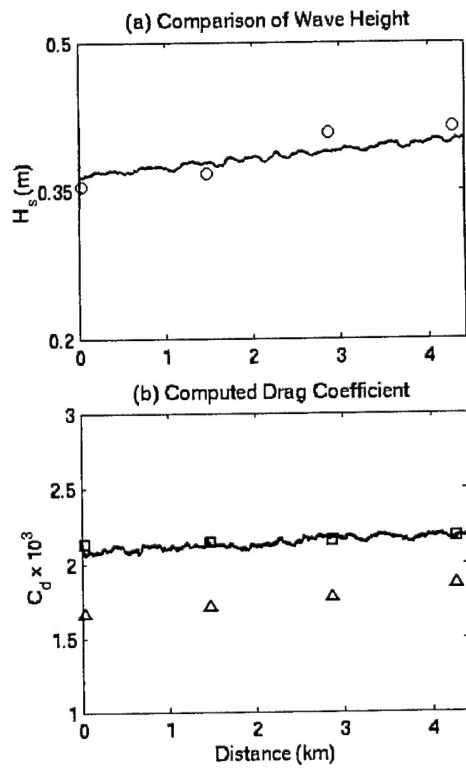


Fig. 4 Comparisons of the computed and measured significant wave heights (top), and the computed and inferred drag coefficients (bottom) in the case of $U_{10} = 10.8 \text{ m/s}$. Solid lines: Boussinesq model results, circles: Young and Verhagen's (1996) observations, squares: Anctil and Donelan (1996), and triangles: Taylor and Yelland (2001).

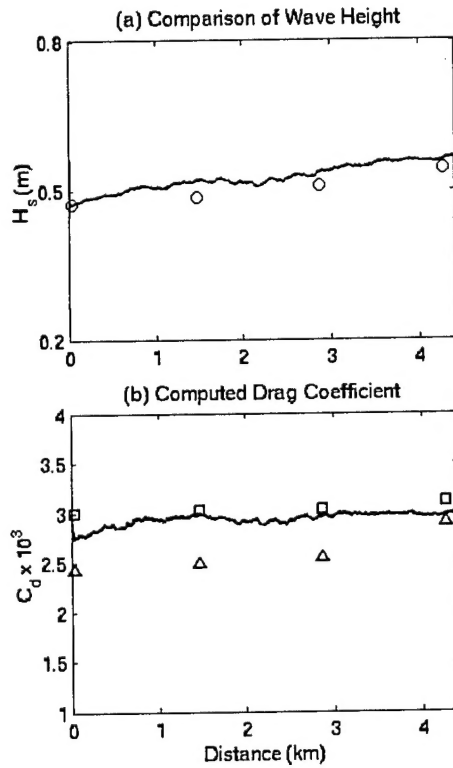


Fig. 5 Comparisons of the computed and measured significant wave heights (a) as well as the computed and inferred drag coefficients (b) in the case of $U_{10} = 15.2\text{m/s}$. Solid lines: Boussinesq model results, circles: Young and Verhagen's (1996) observations, squares: Anctil and Donelan (1996), and triangles: Taylor and Yelland (2001).